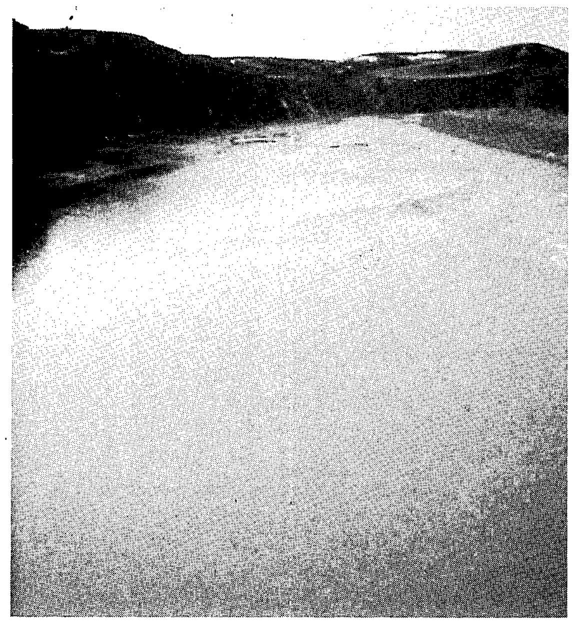
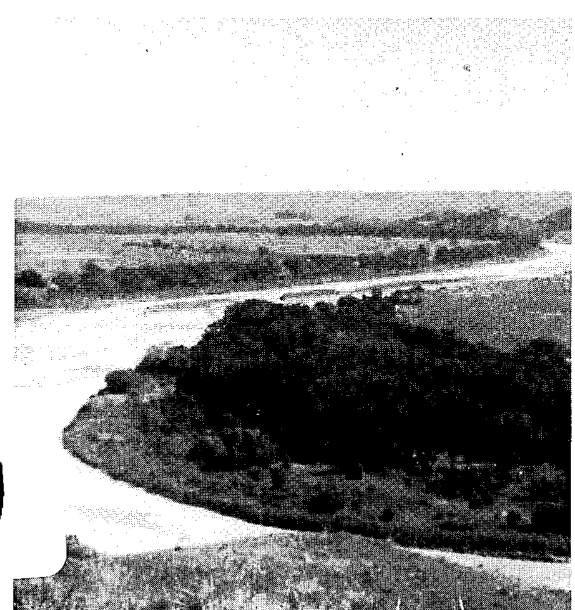
CRREL REPORT 95-15

The Winter Low-Flow Balance of the Semiarid White River, Nebraska and South Dakota

Michael G. Ferrick, Nathan D. Mulherin and Darryl J. Calkins

July 1995





Abstract
Low-flow studies are needed to quantify the effects of water consumption on stream flow, water quality, groundwater resources, and contaminant transport. The low-flow water balance of a river in a cold region is simplified in winter because evapotranspiration is negligible, irrigation water withdrawals and diversions are halted, and precipitation occurs largely as snow, minimizing the spatial and temporal variability of runoff. We investigated the monthly low-flow water balance of White River (Neb. and S. Dak.) reaches over seven consecutive winters. Water going into or out of storage as ice or melt, obtained with an air temperature index model, can be a dominant component of the water balance. The point estimate method is used to account for parameter uncertainty and variability, providing the mean, variance, and limits of dependent variables such as water storage as ice and inflow from a subbasin. Negative surface water yield from several-thousand-square-kilometer subbasins occurred regularly through the period, indicating a significant flow from the river to the alluvial aquifers. The winter water balance results suggest either a perched river or a coupled surface water–groundwater hydrologic system in particular subbasins, consistent with the field investigations of Rothrock (1942). The winter flow exchange between the surface and subsurface can be used to estimate the annual exchange for both hydrologic conditions.

Cover: *Views of the White River at several locations in the basin. Clockwise from lower left: at Crawford, at Oglala, upstream of Kadoka, and near Oacoma. The river decreases in size between Crawford and Oglala even though the basin size increases by a factor of 6. The river increases in size downstream of Oglala to Kadoka, and then maintains its size with a large inflow from the Little White River, downstream to Oacoma. (Photos by N. Mulherin and D. Calkins.)*

For conversion of SI units to non-SI units of measurement consult ASTM Standard E380-93, *Standard Practice for Use of the International System of Units,* published by the American Society for Testing and Materials, 1916 Race St., Philadelphia, Pa. 19103.

CRREL Report 95-15

US Army Corps of Engineers
Cold Regions Research & Engineering Laboratory

The Winter Low-Flow Balance of the Semiarid White River, Nebraska and South Dakota

Michael G. Ferrick, Nathan D. Mulherin and Darryl J. Calkins

July 1995

Prepared for
OFFICE OF THE CHIEF OF ENGINEERS

PREFACE

This report was prepared by Michael G. Ferrick, Research Hydrologist, Nathan D. Mulherin, Research Physical Scientist, Snow and Ice Division, and Darryl J. Calkins, Chief, Geological Sciences Division, Research and Engineering Directorate, U.S. Army Cold Regions Research and Engineering Laboratory (CRREL), Hanover, New Hampshire.

Funding for the work was provided by the U.S. Army Corps of Engineers Civil Works Research Program, *Water Resources of Cold Regions*.

This report was technically reviewed by Edward Chacho of CRREL, Alaska, Robert Ettema of the University of Iowa, and two anonymous reviewers.

Michael Burr, Darwin Rahder, Ralph Teller, and Marvin Stevens of the U.S. Geological Survey (USGS) in South Dakota and James Vosses of the USGS in Nebraska provided data and references that were essential to the completion of this study. Important contributions from members of CRREL included those of Nicholas Goodman (model computations), Matthew Pacillo (figure preparation), Donna Harp (equation and table preparation), and Maria Bergstad (editing).

CONTENTS

ILLUSTRATIONS

Figure

TABLES

Table

The Winter Low-Flow Balance of the Semiarid White River, Nebraska and South Dakota

MICHAEL G. FERRICK, NATHAN D. MULHERIN AND DARRYL J. CALKINS

INTRODUCTION

Water resource development in semiarid regions can lead to declining groundwater levels and streamflow in valleys with permeable soils and interconnected surface–subsurface flow systems, without a corresponding decrease in precipitation. Evapotranspiration losses, water withdrawals, and irrigation return flow affect both the stream and shallow groundwater levels. Soil and geologic characteristics determine the total near-surface aquifer storage, the groundwater recharge from precipitation, and the locations and rates of exchange between groundwater and surface water. The simple characterization of basins by drainage area and precipitation is useful only in homogeneous basins. Subbasin yields at times of low flow can vary widely within a relatively small basin as a result of diverse water-bearing properties of underlying soils and rocks (Schneider 1965, Gerard 1981). Riggs (1972) suggested low-flow discharge measurements at several locations along a stream to define the base flows and the hydrologic homogeneity or heterogeneity of a basin. For a given basin, the annual precipitation and its distribution in time and space determine the quantity of water in storage, and the air temperature regime affects the rate of water loss through evapotranspiration and the storage of water as snow and ice.

Planners need tools for evaluating the effect of proposed changes in water usage in a basin on river flow, potential aquifer yields, water quality, contaminant migration, and other issues (ASCE 1980). In regions with water shortages, an improved understanding of the effects of water consumption is especially important. During periods of low streamflow, the surface water is generally of groundwater origin, but groundwater recharge and discharge are difficult hydrological parameters to quantify. Sophocleous and Perkins (1993) developed a coupled stream–aquifer model with an annual time step and applied it to bound the hydrologic budget imbalance resulting from irrigation development in Kansas. Lacher et al. (1994) measured streamflow, evaporation rates, soil conductivities, pumping rates, and well hydrographs to estimate the rate of aquifer recharge from the Santa Cruz River in Arizona. Abdulrazzak and Sorman (1994) used a water balance approach to estimate flood water losses from ephemeral streams in arid regions, but large spatial and temporal parameter variability introduced uncertainty in the results.

Low flows typically occur in the same season each year. Late summer and winter are low-flow periods in the northern United States and southern Canada (Melloh 1990, Rogers and Armbruster 1990, Wuebben et al. 1992). Kuusisto (1986) reported mean winter-to-summer low-flow ratios that decrease significantly with distance north in Finland. Winter has been generally considered a hydrologically dormant period and has not been extensively studied. However, the exchange of water between a river and its near-surface aquifers is most readily quantified during the winter months. The winter low-flow water balance is simplified because there is negligible evapotranspiration, irrigation water withdrawals and diversions are halted, and precipitation occurs largely in the form of snow, minimizing the spatial and temporal variability of runoff. A complication is that the ice produced in the river can be a large component of the water balance for semiarid basins in even moderately cold regions.

The White River in Nebraska and South Dakota, an uncontrolled tributary of the Missouri River, has a basin of 26,400 km^2, but typical winter monthly average flows are less than 4 m^3/s. The White River basin is heterogeneous, but flows throughout are low and stable in winter. In this report we investigate the winter water balance of the White River in eight subbasins. The water balance is written as a monthly average for river reaches bounded by flow gages. The flow contributions from subbasins and the water storage in the river due to ice production are computed for a series of seven winters, from November 1974 to February 1981. Water going into or out of storage as ice or melt is calculated from an air temperature index model. The point estimate method (PEM) of Rosenblueth (1975) (Appendix A) allows us to apply deterministic relations for ice growth or melt, water storage as ice, and subbasin water yield to the river while still accounting for uncertain or variable parameters in the calculations and flow measurements. The PEM provides an expected value, variance, and estimated limits of the probability distributions that characterize the dependent variable in each of these calculations. Our objective is to quantify the water yield from each subbasin to the river and, where possible, the river–subsurface flow exchange.

GENERAL HYDROLOGY OF THE BASIN

The White River basin lies in an unglaciated part of the Missouri Plateau characterized by undulating uplands and wide floodplains along the larger streams. The location and basin maps for the White River given in Figure 1 indicate the basin boundaries, the primary tributaries, the Little White River basin boundaries, and the U.S. Geological Survey (USGS) stream gaging stations and meteorological data stations used in this study. The streamflow gages on the main river and the Little White River are listed in Table 1, along with the drainage area of each nested subbasin, the annual and winter average discharges for the period of record, the gage datum, the approximate river location of each gage or length of the reach between gages, the average channel slope, and the linear distance between gages. The winter average discharge was obtained from November through February monthly averages. The channels of the White and Little White Rivers are highly mobile within the floodplain, and river location is not published by the USGS. The approximate

Table 1. Basin, flow, and river parameters corresponding to USGS gages on the White and Little White Rivers.

Location	*Drainage area (km^2)*	*Avg. discharge (m^3/s)*			*Gage datum or elev. (m)*	*Gage location or reach length (km)*	*Approx. river slope*	*Linear distance (km)*	*River/ linear distance ratio*
		Annual	*Winter*	*Ratio*					
White River									
Crawford	810	0.57	0.59	1.04	1115.5	804.0			
Cr–Og	4,890	0.96	–0.06	—		196.0	0.0013	78.9	2.48
Oglala	5,700	1.53	0.53	0.35	869.8	608.0			
Og–Ka	7,300	6.03	1.21	0.20		260.0	0.00086	118.5	2.19
Kadoka	13,000	7.56	1.74	0.23	646.8	348.0			
Little White River									
Martin	800	0.54	0.38	0.71	928.1	196.0			
Ma–Ve	700	0.99	0.69	0.70		62.0	0.0013	33.2	1.87
Vetal	1,500	1.53	1.07	0.70	847.6	134.0			
Ve–Rb	1,140	1.61	1.40	0.87		65.6	0.0022	37.6	1.74
Rosebud	2,640	3.14	2.47	0.79	699.5	68.4			
Rb–WR	1,430	0.48	0.17	0.35		45.1	0.0026	32.5	1.39
White River	4,070	3.62	2.64	0.73	583.0	23.3			
WR–Confl					541.0	23.3	0.0018		
White River									
Ka+WR	17,070	11.2	4.37	0.39					
KaWR–Oa	9,330	3.7	–0.90	—					
Ka–Oa						342.0	0.00066	158.0	2.16
Oacoma	26,400	14.9	3.47	0.23	419.8	6.0			

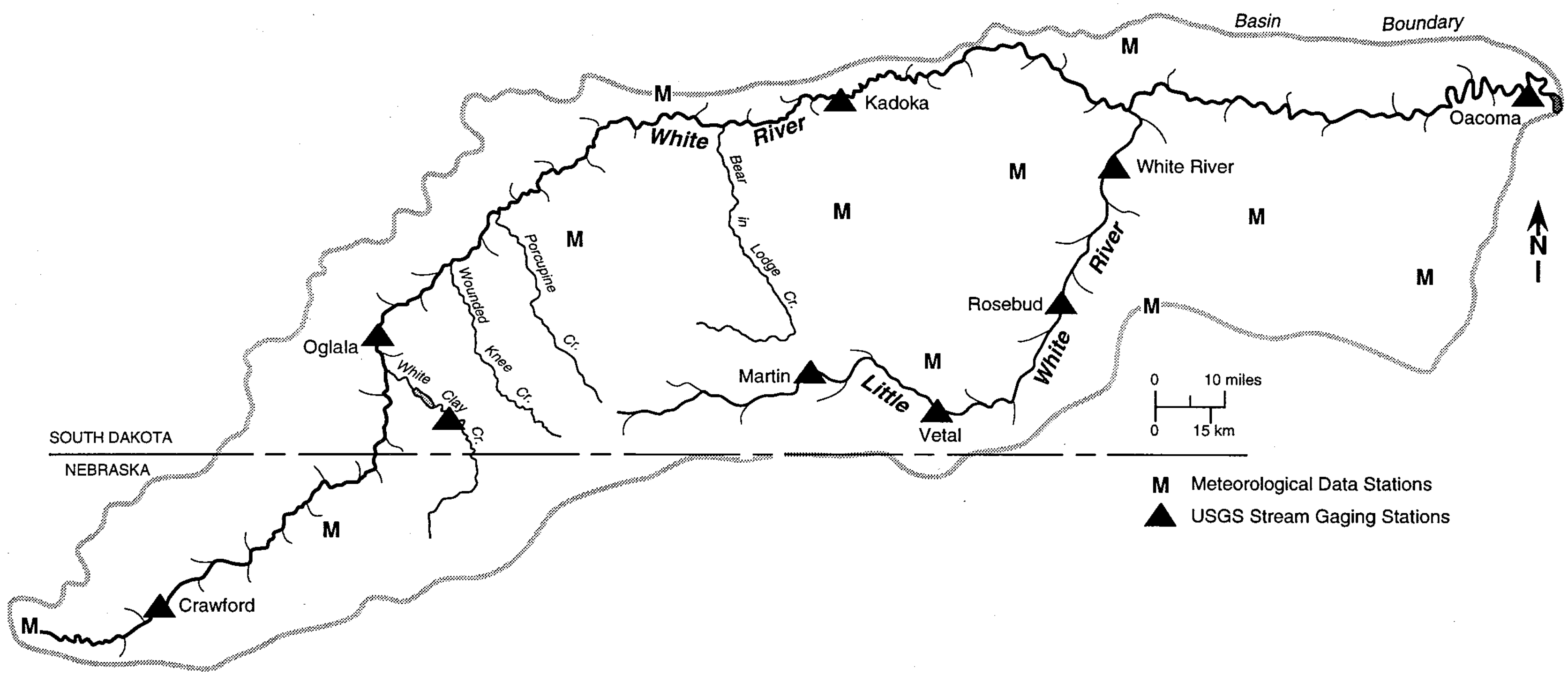


Figure 1. White River basin and location maps.

river locations in Table 1 were obtained from recent maps using a map wheel. The White River is generally more sinuous than the Little White River, with generally higher river/linear distance ratios. The Little White River below Martin has an average slope of 0.0020, more than twice the 0.00087 of the White River below Crawford.

Sando (1991) detailed the surface water diversions and groundwater withdrawals for irrigation over the last 50 years in the White River basin upstream of the South Dakota state line. Water use is zero for November through April but currently averages 50% of the flow at the state line gage for the remainder of the year. All tributaries in the basin above the state line are ephemeral (Sando 1991) and, more generally, streams in the White River basin are unreliable sources of water (Ellis and Adolphson 1971). Most streamflow occurs in response to precipitation and runoff during spring and early summer, with a large component of the March flow due to snowmelt. Rothrock (1942) reported extensive field investigations throughout the river valley over a 43-km reach from Kadoka downstream. The alluvium there is between 7.5 and 12.5 m deep, composed mostly of sands and gravels, overlying Pierre shale. Near-surface aquifers in the alluvial deposits are permeable and readily exchange water with the river. The groundwater level is below the river surface during dry seasons, rising to river level during wet seasons. When river flow ceases, deep pools remain wet, reflecting the water table level. Short reaches of the perennial Bear-in-the-Lodge, Porcupine, Wounded Knee, and White Clay Creeks intercept the water table, allowing groundwater inflow to supplement runoff. The larger tributaries supply groundwater with a different chemical signature directly to the alluvium of the White River. The wind-blown sand deposits in the Little White River basin above Rosebud are permeable, minimizing surface runoff and providing more consistent flows than are found elsewhere in the basin (Ellis et al. 1971).

The subbasin above the Crawford gage provides stable river flows throughout the year. Significant groundwater input to the river and minimal surface runoff cause this stability, which is interrupted only occasionally by large spring and summer events. The annual hydrograph of monthly average flows at Oglala has peaks in both March and June and low flows in the fall and winter. The average winter flow at Oglala is less than that at Crawford (Table 1), even though the basin is seven times larger. Rothrock (1942) reported that "in the upper part of the valley" a considerable flow will frequently disappear within 35 km due to groundwater recharge. The annual hydrograph of monthly average flows at Kadoka also has double peaks and the same general shape as that at Oglala, with flow decreasing through the fall and into midwinter. However, the increase in discharge between these gages is generally significant, due largely to flow contributions of the perennial creeks. The yield of the Little White River basin as surface water is indicated by gage White River (WR) at 23 km above the White River confluence. Spring flows on the Little White are high and variable, while fall and winter flows are lower and more stable. Summer flows are occasionally high, but generally consistent with the groundwater-inflow-dominated fall and winter conditions. The annual hydrographs of monthly average flow for all gages in the Little White basin have single peaks in either March or April. The monthly average White River flow at the Oacoma gage near the mouth can vary dramatically between seasons, especially spring and summer, and years. Winter flows are more consistent and extremely low by comparison, with 0.6 m^3/s at Crawford and 2.6 m^3/s from the Little White River equaling almost the entire flow at Oacoma from subbasins representing only 18% of the total drainage area.

Figure 2 provides a breakdown of annual and winter average water yields to the river of sequential subbasins of the White River basin, delineated by the primary streamflow gages. The highest annual and winter water yields in the entire basin occur in a subbasin of the Little White River between Martin and Rosebud. This subbasin has an annual water yield 2.5 times greater, and a winter yield 8.7 times greater, than that of the complete basin. The ratio of winter to annual average discharge is greater than 1 for the subbasin above Crawford and greater than 0.7 for all subbasins of the Little White River above the gage at Rosebud. However, for most of the main-stem White River and the Little White River below Rosebud, the winter flows are only 23 to 35% of the annual average. The subbasin between Crawford and Oglala has the lowest annual yield in the entire basin and a negative winter yield. Other subbasins with low yields both annually and in winter are the Little White below Rosebud and the White below Kadoka. Unlike the other main-stem subbasins below Crawford, the subbasin between Oglala and Kadoka has a high annual yield and a positive winter yield. The

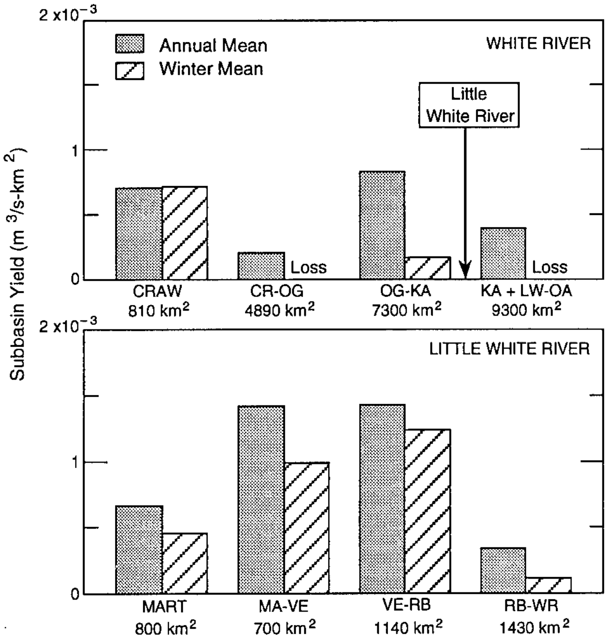


Figure 2. Annual and winter average water yields for the period of record in sequential subbasins of the White River basin.

hydrogeologic maps of Ellis and Adolphson (1971) and Ellis et al. (1971) support the conclusion that the primary reason for the widely differing yields of the White River subbasins is differences in near-surface geology.

RIVER ICE GROWTH AND MELT

The extreme low flows in winter on the main stem of the White River greatly increase the importance of water storage as ice in the water balance. We now develop a method to quantify the storage of water as river ice and its release as melt. A large number of processes occur during ice formation and growth in rivers that contribute to variable ice thickness. For example, rapids can remain open all winter, producing ice at a high rate, while slower and deeper reaches situated downstream can collect this ice and develop thick deposits. However, we are interested in characterizing ice growth at a monthly time scale over long reaches of a shallow, mildly sloped river. Therefore, we assume that these local, mechanically induced variations in thickness occur about a mean that is dictated by the air temperature regime. The physically based air temperature index model of ice growth given by Ashton (1989) is applied:

$$\frac{dh}{dt} = \frac{1}{\rho_i L} \frac{(T_m - T_a)}{\left(\frac{h}{k} + \frac{1}{H_{ia}} + \frac{h_s}{k_s}\right)} \qquad (1)$$

where h and h_s are thicknesses (m) of the ice and its snow cover, t is time (s), k and k_s are thermal conductivities of the ice and snow (W/m °C), L is the latent heat of fusion (J/kg), ρ_i is the density of ice (kg/m^3), H_{ia} is the ice–air or snow–air heat transfer coefficient (W/m^2 °C), and T_m and T_a are the ice melting point and air temperatures (°C), respectively. Integrating eq 1 and taking the positive root of the resulting quadratic equation, we obtain the final or end-of-the-month ice thickness h_f as

$$h_f = \left[(h_i + h_r)^2 - \frac{2kT_a \Delta t}{\rho_i L}\right]^{1/2} - h_r \qquad (2)$$

where

$$h_r = k\left(\frac{1}{H_{ia}} + \frac{h_s}{k_s}\right),$$

h_i is a given initial or start-of-the-month ice thickness, and Δt is the monthly time increment (s). The parameter h_r in eq 2 represents an equivalent ice thickness corresponding to the thermal resistance of the snow and the interface with the air.

The melting of a river ice cover can occur on both the top and bottom surfaces. The heat supplied by the water to the bottom surface becomes dominant during ice breakup in the spring as the area of open water upstream becomes large (Prowse 1990). However, in the early stages of ice melt when most of the river is still ice covered, the direct heat flux from the atmosphere to the top surface is dominant. We assume that ice melt during the winter period represents an "early stage" of melt and that the snow on the ice at this time is negligible. These assumptions allow us to use a simplified form of eq 1 to estimate the melt, and integration yields

$$h_f = h_i - \frac{H_{ia}}{\rho_i L} T_a \Delta t \ . \qquad (3)$$

With h_f known from either eq 2 or 3, we can obtain the monthly average flow Q_{ice} that has gone into or out of storage as ice or melt:

$$Q_{ice} = \frac{\Delta x}{\Delta t} \frac{(B_{in} + B_{out})}{2} (h_f - h_i) \quad (4)$$

where Δx is the reach length (m) and B is channel width at the upstream (*in*) and downstream (*out*) ends of the reach. Equation 4 assumes that the average width of the river in a reach can be obtained by averaging the widths at each end.

WINTER RIVER WATER BALANCE

Over the period of record, the annual and winter water yields to the White River indicated a wide range of subbasin hydrologic conditions. In particular, major differences in subbasin yields to the river were evident in the winter. We now develop a monthly winter water balance for a river reach that includes variable water storage in the river channel with large changes in flow, the formation or melt of river ice, and flow exchange with the corresponding subbasin. The effects of unsteadiness on the water balance are negligible during low-flow periods and will be neglected. The net inflow to the river from a subbasin Q_{sub} has tributary and groundwater components:

$$Q_{sub} = Q_{gw} + Q_{t1} + Q_{t2} + Q_{t3} + \dots \quad (5)$$

where Q_{gw} is groundwater inflow and Q_{t1}, Q_{t2}, and Q_{t3} are tributary inflows. The net groundwater exchange between adjacent subbasins is needed to determine if Q_{gw} is supplied by the subbasin or if it contains a significant intersubbasin component. The groundwater flow in the alluvium Q_{int} at a subbasin boundary is given by Darcy's Law as

$$Q_{int} = AKJ \quad (6)$$

where A is cross-sectional area, K is hydraulic conductivity of the alluvium, and J is the hydraulic gradient. Rothrock (1942) obtained the data needed to evaluate the downvalley groundwater flow near Kadoka as $Q_{int} = 0.017\ m^3/s$. The net groundwater exchange of a subbasin with its neighbors is the difference between Q_{int} values at the bounding stream gages. The downvalley groundwater flow along the main-stem White River is small relative to the subbasin flows given in Table 1, and the net exchange is probably even smaller. Therefore, we will assume that Q_{gw} is supplied by the subbasin.

The flow storage in the channel Q_{st} caused by significant changes in the monthly average flow can be computed for a river reach as

$$Q_{st} = \frac{\Delta x}{\Delta t} \left[\frac{B_{in}\Delta Y_{in} + B_{out}\Delta Y_{out}}{2} \right] \quad (7)$$

where ΔY (m) is the channel depth change at the upstream (*in*) and downstream (*out*) ends of the reach, assuming that depth change can be adequately described by averaging the end values. The depth changes can be determined from the measured average discharge at each gage for the present and previous months and corresponding river stage data.

The winter water balance for a river reach delineated by a pair of stream gages is depicted in Figure 3 and written as

$$Q_{in} + Q_{sub} - Q_{ice} - Q_{st} - Q_{out} = 0 \quad (8)$$

where Q_{in} and Q_{out} are the flows measured at the upstream and downstream gages, respectively. In low-flow months the subbasin flow exchange may be almost exclusively with the groundwater. As tributary inflows are always nonnegative, $Q_{sub} < 0$ implies groundwater recharge from the river.

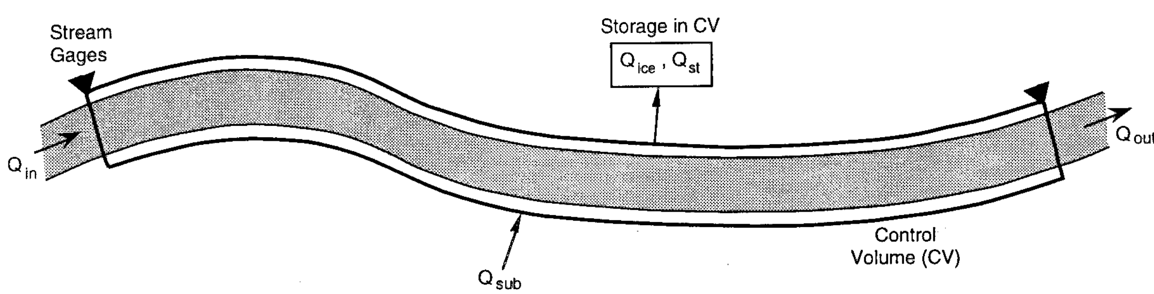


Figure 3. Schematic diagram of the control volume used to obtain the winter water balance for a river reach.

All of these computations are nested, with changes in ice thickness obtained for a given month and then used to find Q_{ice}. With Q_{st} and Q_{ice} known from eq 7 and 4, and Q_{in} and Q_{out} known from gage records, we can obtain Q_{sub} with eq 8. Finally, tributary inflows are used with Q_{sub} to obtain the groundwater discharge or recharge Q_{gw} from eq 5.

POINT ESTIMATE METHOD

Mean values can now be determined for Q_{ice}, Q_{st}, Q_{sub}, and Q_{gw}, but they would not account for the known variability of input parameters such as air temperature, heat transfer coefficient, and channel width. In addition, many of the measured or estimated parameters contain uncertainty that contributes to the uncertainty of the corresponding dependent variable. We will use the Rosenblueth (1975) point estimate method (PEM) to account for and quantify the uncertainty in our deterministic winter water balance. The independent variables in each deterministic equation that contain uncertainty are considered random variables. The first two or three moments of each random variable and the correlation coefficients between variables are given as input, quantifying the variability or uncertainty. The PEM provides the mean, variance, and limits of the dependent variable, uniquely specifying a beta distribution (Harr 1977) that describes the uncertainty of a function of random variables. The estimated mean value is equivalent to a second-order Taylor series approximation, and the variance is a first-order estimate. The method is algebraic, replacing the distribution of each random variable by point estimates and not requiring the computation of derivatives. The PEM offers several advantages over a deterministic approach. A computed mean value has much greater importance when the variance is small, but variance is unknown in a deterministic model. The random variables contributing most of the uncertainty to the results can be readily identified, which can help to refine data collection. In addition, the interpretation of PEM results is straightforward. For example, though a river reach may have positive inflow from its subbasin based on mean values, there may be a significant probability that the flow direction is the opposite.

We apply the PEM to the ice growth or melt in eq 2 or 3 by considering air temperature, the air–ice heat transfer coefficient, and the initial ice thickness as the primary random variables. The mean of H_{ia} was taken as 20 W/m^2 °C with a standard deviation of 5 W/m^2 °C, representative of the data presented by Ashton (1989). Ice density and thermal conductivity are correlated random variables, but their variability is minor. The mean ice density was taken as 900 kg/m^3 with a standard deviation of 15 (Mulherin et al. 1992). The mean thermal conductivity of ice was taken as 2.17 W/m °C with a standard deviation of 0.1, and the latent heat of fusion was assumed constant at 333,400 J/kg (Ashton 1986). Data were unavailable to quantify the snow depth on the ice. However, because the overflow of water on grounded ice often incorporates the snow into the ice surface, h_s was assumed to be negligible.

In computing Q_{ice} with eq 4, the independent random variables are river distance between the gages, channel width at each gage, and correlated initial and final ice thicknesses. The estimated mean river distances between the gages are given in Table 1. Based on multiple trials, the measurement error in obtaining these distances from maps was about 2% of the distance, and in addition, the movement of the river within the floodplains could alter the distances from those shown on the maps by a few percent of the length. Therefore, we assume a coefficient of variation of 0.05 for reach length. Ice formation causes the flow in the wide, flat channel to shift its location. Ice then freezes to the bed in nearshore and bar areas, and water flow is restricted to only a portion of the apparent width. The channel width of the ice/water interface was measured by the USGS each time a discharge rating was done at a gage. The channel cross section and discharge at which these measurements were made varied. We used all available measurements during ice-covered flow conditions to obtain the mean width and its variance near each gage. The mean width varied from 4 m on the Little White River at Martin to 23 m on the White River at Oacoma. The coefficient of variation of the river width varied between 0.11 and 0.46. These data indicate that the width of the White River at low flow can vary significantly over short distances. Systematic analysis of aerial photographs of the river taken at low flow just prior to ice formation or extensive ground measurements along the river would best quantify the distribution of width.

The Q_{st} computation in eq 7 has reach length and stream width and correlated depth changes at the ends of the reach as independent random variables. We assume that Q_{st} is generally negligible but evaluate it for the pairs of months of largest flow increase and decrease in the period of record. Q_{sub} in eq 8 can be evaluated considering each component of the water balance as a random variable. The measurement error for discharge at the gages in winter is typically about 8%, as reported on the USGS discharge measurement notes. We will use this value as the coefficient of variation for the measured monthly average discharges at the gages. Q_{in} and Q_{out} correlations were computed for each pair of gages over the period of study and have coefficients that increase with distance downstream.

RESULTS

Seven consecutive winters from November 1974 to February 1981 provide a representative range of temperature and hydrologic conditions for analysis. Mean monthly air temperatures are available at the 11 meteorological stations in Figure 1 for the seven-winter study period. To account for the variability indicated by these temperatures, a basin mean temperature and standard deviation were obtained and are presented in Table 2. The coldest winter of the study period was 1978–79, and January 1979 was the coldest month. Five of the six other winters had less than half of the freezing degree-days of this winter. With these temperatures as input, we obtained the ice growth or melt for each month, and the mean thickness and standard deviation at the end of the month are given in Table 2. December 1977 and January 1980 were the months of maximum ice growth, and February 1976 was the month of maximum melt in this period of record.

Figures 4 and 5 give results for January 1979, the coldest month of the study. The mean, standard deviation, and corresponding beta distribution for Q_{ice} and Q_{sub} of each main-stem White River reach are presented in Figure 4. The storage of water as ice is an important term in the water balance of the White River below Oglala, but it is of less significance farther upstream and on the Little White River, where the stream widths are small and the flows are relatively high. The Crawford–Oglala subbasin had a negative yield to the river. The distributions for Q_{ice} and Q_{sub} have similar shapes in the two reaches below Oglala. There is a small probability that the Oglala–Kadoka subbasin had a negative yield and the Kadoka–Oacoma subbasin had a positive yield to the river. The mean and standard deviation data are repeated in Figure 5 from upstream (left) to downstream (right), together with corresponding flows at the gages. The river flows diminished from Crawford to Kadoka and then recovered somewhat at Oacoma, due to a significant inflow from the Little White River. The mean water storage as ice increased in successive reaches downstream, as did its variance. The variance of subbasin flow exchange with the river also increased in the downstream direction.

Table 2. Mean and standard deviation for monthly air temperature over the basin T_a (°C), and corresponding cumulative ice thickness h_f (m).

Year	*Variable*	*November*	*December*	*January*	*February*
1974–75	T_a	2.8, 1.2	–2.2, 1.1	–4.4, 0.9	–7.2, 0.9
	h_f	0, 0	0.19, 0.072	0.40, 0.056	0.61, 0.047
1975–76	T_a	1.1, 1.0	–1.7, 1.3	–5.0, 1.4	1.7, 1.2
	h_f	0, 0	0.15, 0.096	0.40, 0.074	0.12, 0.23
1976–77	T_a	–0.6, 0.6	–2.8, 0.8	–10.6, 1.3	1.1, 1.2
	h_f	0.06, 0.06	0.26, 0.054	0.63, 0.046	0.45, 0.20
1977–78	T_a	0.6, 2.1	-6.1, 1.2	–12.8, 1.4	–10.0, 1.4
	h_f	0, 0	0.38, 0.052	0.75,0.046	0.93, 0.045
1978–79	T_a	–3.9, 1.0	–8.3, 1.4	–13.9, 0.9	–8.9, 1.8
	h_f	0.28, 0.051	0.58, 0.051	0.89, 0.041	1.0, 0.045
1979–80	T_a	0, 1.7	0.6, 1.3	–6.1, 1.1	–3.3, 1.1
	h_f	0, 0	0, 0	0.38, 0.048	0.49, 0.052
1980–81	T_a	3.9, 1.5	–1.7, 1.0	–0.6, 0.7	–1.7, 0.9
	h_f	0, 0	0.16, 0.073	0.20, 0.080	0.28, 0.076

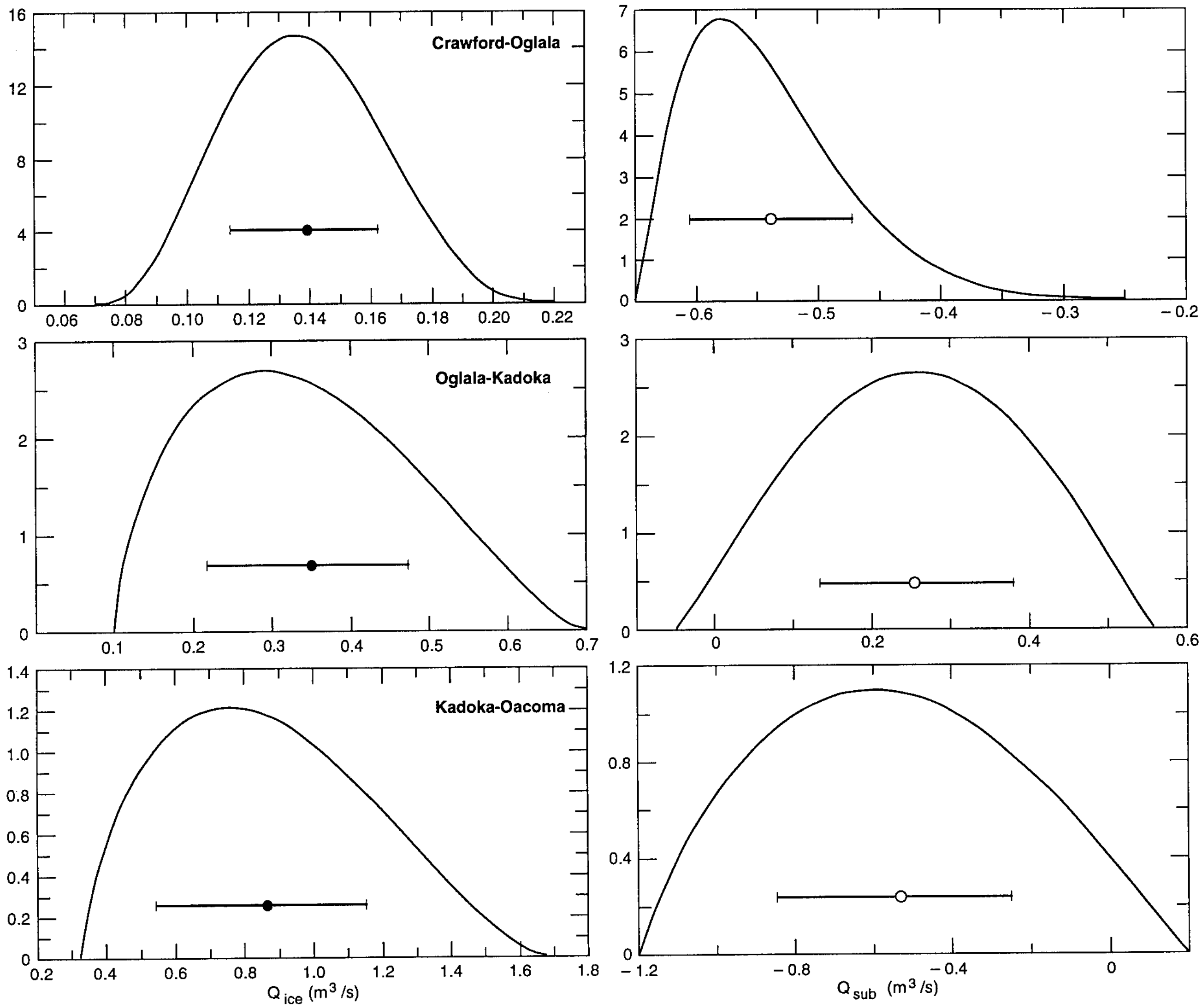


Figure 4. Beta distributions with means and standard deviations indicated for water storage as ice and subbasin inflow in January 1979, by reach between main-stem White River gages. Positive Q_{ice} is water loss due to ice growth, and positive Q_{sub} is flow to the river from the subbasin.

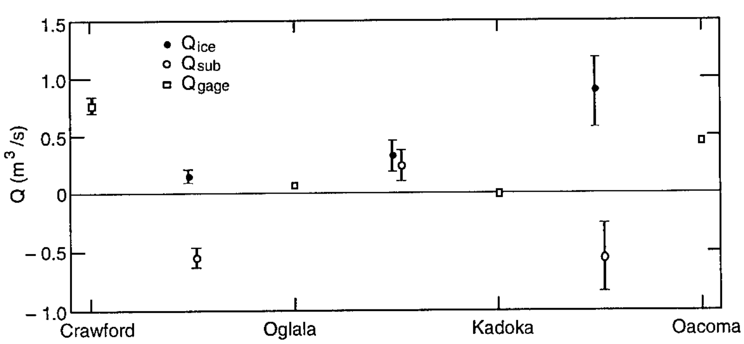


Figure 5. January 1979 monthly flow at main-stem White River gages, water storage as ice, and subbasin inflows of the reaches between gages. Both the mean and standard deviation are given for each parameter.

Results for each main-stem subbasin are detailed in Figure 6 for 1978–79, the coldest winter of the study. The ice production was consistent over the winter except for a drop in February, with increasing mean and variance in the downstream direction. The mean subbasin flow exchanges were consistently negative for Crawford to Oglala and Kadoka to Oacoma, but positive between. The magnitudes of Q_{ice} and Q_{sub} are generally comparable to or larger than the river flows at the gages for this entire winter. Figure 7 gives corresponding results for 1979–80, another low-flow winter with only average air temperatures. The flow storage due to ice growth was reduced somewhat compared with 1978–79, but subbasin yields to the river were generally comparable. Water storage as ice must be considered in a winter water balance for semiarid regions, but extreme cold may not significantly alter basin hydrologic response. In both winters the magnitudes of Q_{sub} obtained for all the main-stem subbasins are large compared to Q_{int} calculated by Rothrock (1942), in support of our assumption that the net groundwater exchange between subbasins is negligible.

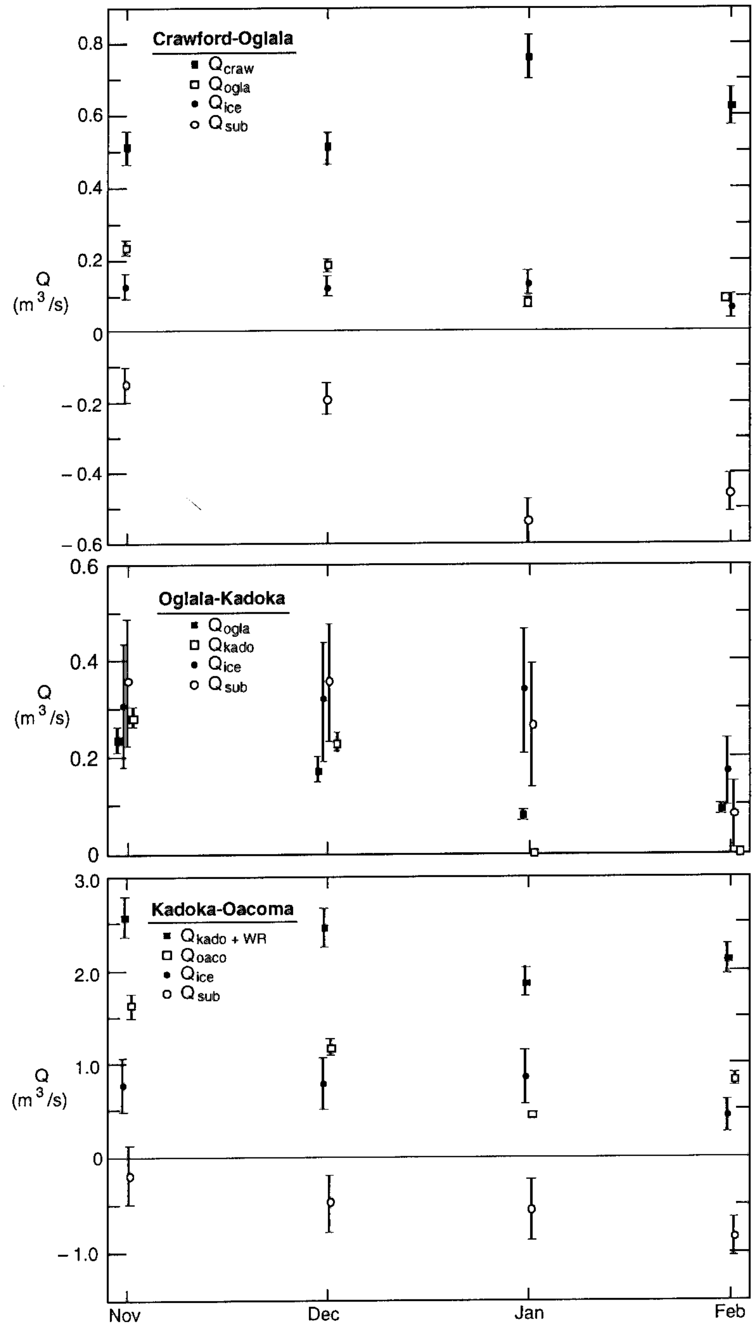


Figure 6. Winter 1978-79 monthly flows at main-stem White River gages, water storage as ice, and subbasin inflow for the reaches between the gages.

The Crawford–Oglala subbasin had only 3 of 28 months with a net inflow to the river: February 1976, the month of maximum ice melt in the period with relatively high flows indicating runoff throughout the basin, and two months in the 1979–80 winter. Excluding the four highest inflow months, the mean Q_{sub} for the period was –0.318 m^3/s with a standard deviation of 0.125 m^3/s. The only perennial stream in the reach is White Clay Creek. To obtain Q_{gw} the analysis was repeated with Crawford plus White Clay Creek providing Q_{in}. With this change the correlation between reach inflow and outflow increased from 0.409 to 0.721. The mean Q_{gw} obtained was –0.483 m^3/s, with a standard deviation of 0.091 m^3/s. Groundwater recharge occurs consistently at a steady rate during winter and represents a significant flow loss from the river.

The Oglala–Kadoka subbasin produced consistently positive inflows. The three large inflow months were the melt in February 1976 and prefreezeup combined with relatively high precipitation in November 1977 and 1979. The months of small local inflows were those with low flows at both gages and minor ice production. Excluding the four highest inflow months, the overall mean Q_{sub} was 0.39 m^3/s, with a standard devia-

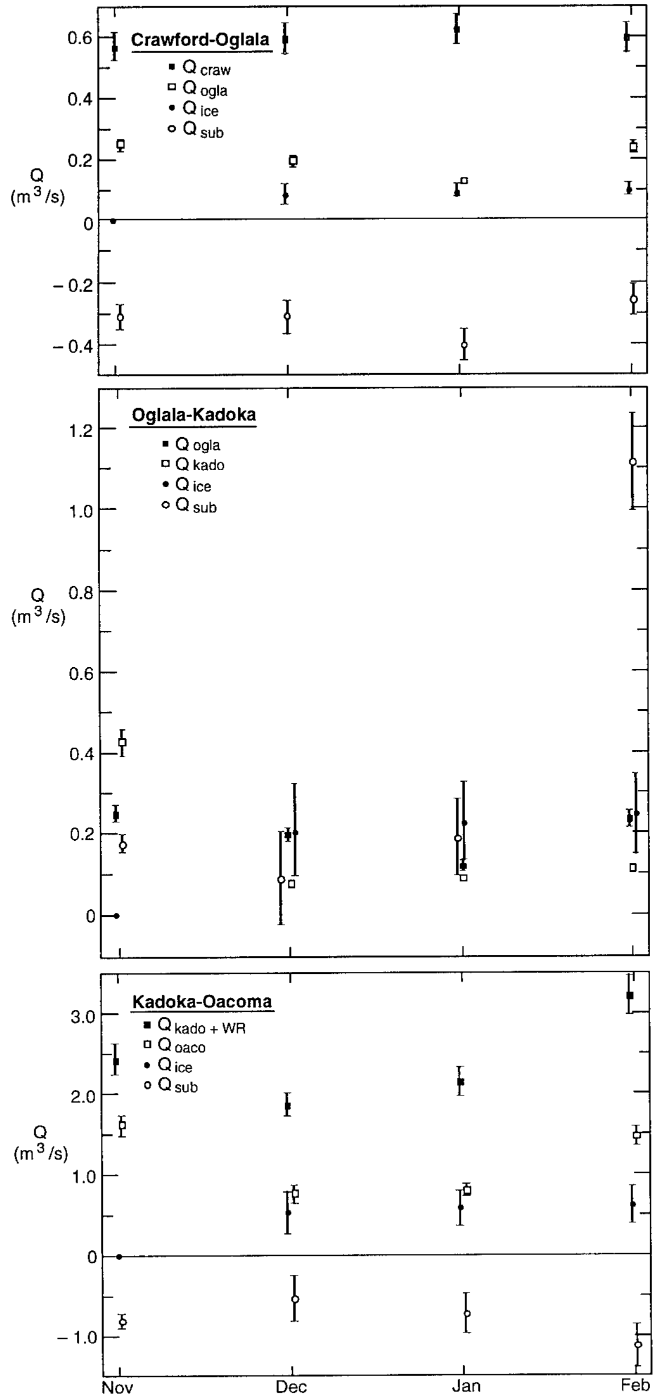


Figure 7. Winter 1974–75 monthly flows at main-stem White River gages, water storage as ice, and subbasin inflow for the reaches between the gages.

tion of 0.32 m^3/s and all positive monthly means. The river–groundwater exchange in this reach is clearly different from that in the adjacent subbasin upstream, but the quantity and direction are masked by inflow from three perennial creeks. Two of these creeks were gaged in 1992–93, having a combined winter flow 1.7 times that of White Clay Creek. We use two times the White Clay Creek flow as a conservative estimate of tributary inflow to this reach. The overall mean Q_{gw} was obtained, again excluding the four high inflow months, as 0.11 m^3/s with a standard deviation of 0.31 m^3/s and 12 negative mean inflow months. The river–groundwater exchange in either direction near Kadoka is consistent with Rothrock (1942), but a more quantitative understanding will require additional tributary flow data.

The monthly subbasin inflows on the Little White River above Rosebud are always positive. The mean inflows downstream of the Rosebud gage are usually positive, but there is a significant probability of negative inflows for almost all months of the study period. These significantly reduced subbasin yields to the river represent a transitional behavior between the high yields of the Little White basin upstream and the low yields of the White River subbasin immediately downstream. The Little White River inflow to the Kadoka–Oacoma reach is typically much greater in winter than the White River main stem flow at Kadoka. Significant positive subbasin inflows to this furthermost downstream White River reach were computed for February 1976, November 1977, and November 1979, the same months as the adjacent subbasin upstream. However, the local monthly subbasin inflows are again generally negative in the winter. Flow losses in 7 of 28 months exceeded 1.1 m^3/s, with a maximum loss of 2.5 m^3/s. Neither the coldest months nor those with maximum ice growth correspond to months of maximum flow loss from this reach.

We now consider the assumption of negligible channel storage in the water balance. For most of the seven study winters, the change in average flow of consecutive months was small throughout the basin, and Q_{st} could be neglected without significant error. The largest flow increase of the study period on the main-stem White River occurred between January 1976, a cold, low-flow month, and February 1976, a month of significant melt, runoff, and relatively high flow. The consecutive monthly flows at Oglala and Kadoka increased by factors of 8 and 29, respectively. We obtained Q_{st} using USGS gage rating data from these months. The mean subbasin inflow to the river increased from 9.55 to 9.71 m^3/s when channel storage was conchange of less than 2%. Similarly, a flow decrease by a factor of 20 occurred at Kadoka between October and November 1980. Mean subbasin inflow decreased from 0.24 to 0.02 m^3/s with channel storage included in the water balance. The direction of Q_{gw} can be reversed by the channel storage term for decreasing flow conditions.

HYDROLOGIC IMPLICATIONS

The winter hydrologic balance is useful for determining flow quantities and directions, but insights on the annual exchanges would be very important. The balance between groundwater consumption and recharge determines the long-term availability of the resource. Conversely, increased flow losses from the river affect water quality, aquatic habitat, and surface water availability. Figure 8 depicts the possibilities for a near-surface aquifer that is recharged by the river during winter. The first has the river perched above the water table throughout the year, resulting in a continuous loss of water from the river by unsaturated flow. For given alluvial bed conditions the flow loss would be proportional to the wetted perimeter and depth of the river, each generally increasing with river flow. The other possible condition is a coupled stream–aquifer system, with a fluctuating relationship between river and groundwater levels throughout the year. In the case depicted, a summer of groundwater withdrawals and insufficient recharge has caused the water table to fall to a minimum by the end of October. The end of the irrigation season, together with recharge from the river over the winter, causes a recovery of the levels. By April the water table is at river level, and it continues to rise through June. After that time groundwater discharge and withdrawals cause the levels to fall, reaching river level in July and continuing down with sustained withdrawals toward the fall minimum. A plot of flow loss from the perched river is depicted in Figure 9, together with the mean annual hydrograph of the White River at Oglala. With a perched river, the winter flow loss extrapolated through the year would provide a lower-bound estimate of annual groundwater recharge from the river. The case of a coupled hydrologic system is depicted in Figure 10. "Net groundwater loss" in this figure is the withdrawal restricted to the period with the water table at or below river level. As drawn, the flow loss in winter is nearly equal to the annual river flow loss. An extrapolation of the winter losses provides an upper bound for recharge from the river.

The net yield per square kilometer of a subbasin or the net flow loss per square meter of channel bed of three main-stem reaches are given in Table 3 for relatively dry months (January 1977 and 1979) and wet months (February 1976 and November 1977). Tributary inflows considered apart from

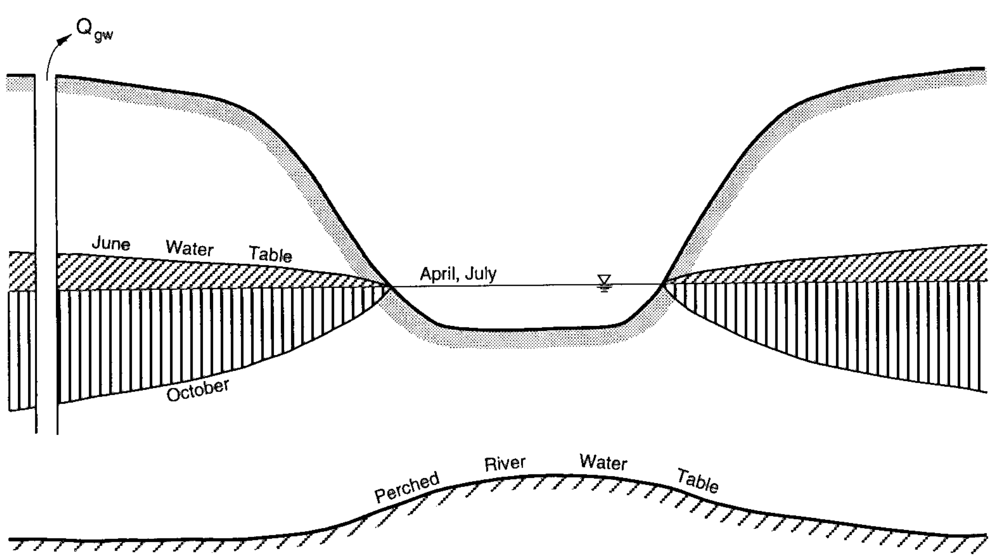


Figure 8. Schematic diagram of a river that recharges the groundwater during winter. Groundwater withdrawal is indicated. The river may be perched above the water table or directly connected to the water table in a coupled hydrologic system. Water tables in the diagonal shading discharge to the river, and those in the vertical shading or below are recharged by the river.

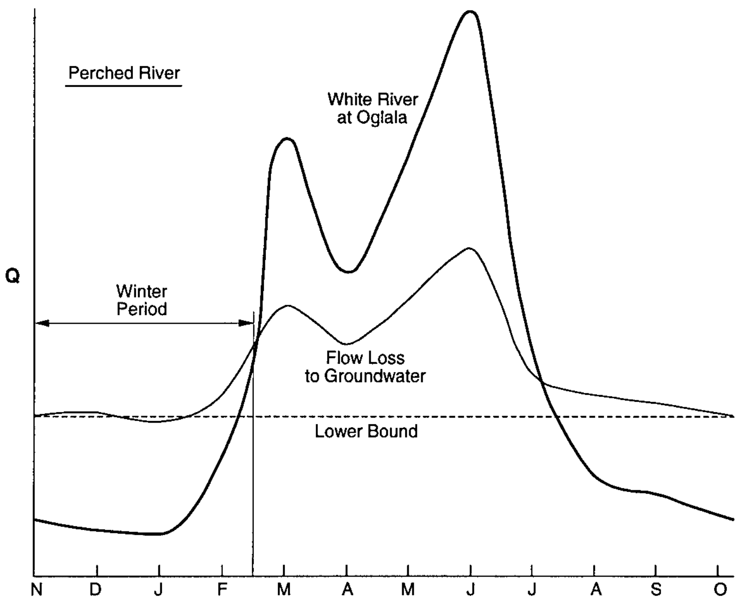


Figure 9. Hypothetical flow loss from the river to the groundwater for a perched river. The mean annual hydrograph for the White River at Oglala is used as a reference. The flow loss during winter extended over the year provides a lower bound for the annual flow loss.

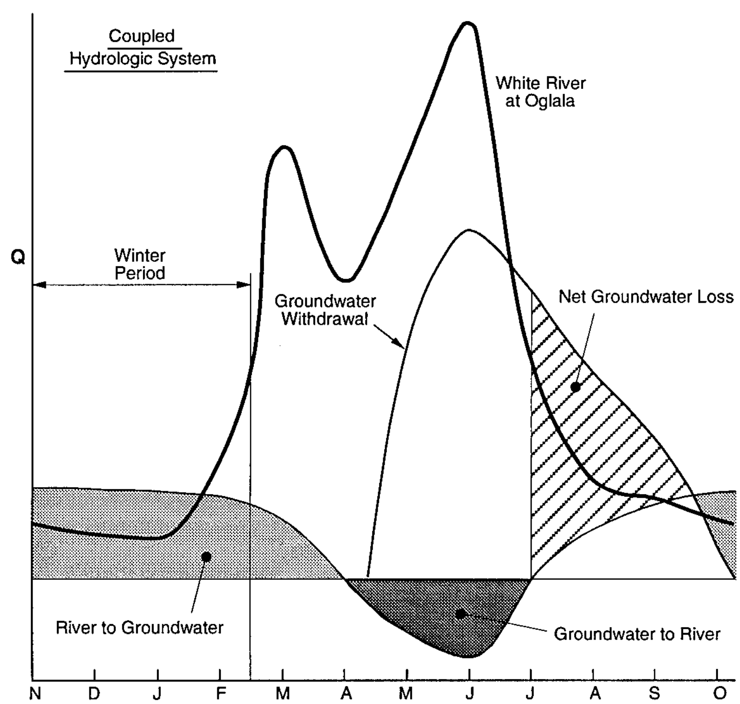


Figure 10. Hypothetical flow exchange between the river and the groundwater for a coupled hydrologic system. The mean annual hydrograph of the White River at Oglala is used as a reference. The curve of groundwater withdrawal is based on irrigation data of Sando (1991) and represents losses above recharge from precipitation and return flow. The lighter shading represents volumetric loss from the river, in balance with net groundwater loss (diagonal shading), and the darker shading represents volumetric groundwater to river exchange.

the remainder of the subbasin were White Clay Creek (WCC), two times White Clay Creek (2WCC), and the Little White River at WR. In a given month with no additional tributary inflows, the subbasin yield or river flow loss given in Table 3 is the groundwater exchange. With few exceptions the Crawford–Oglala reach had consistent flow losses over the channel area of between 4.0 and 4.9×10^{-7} $m^3/s\text{-}m^2$, a midrange seepage velocity for fine sands and silts (Bear 1972). The evidence supports the hypothesis of a predominantly perched river through this reach. Small, variable flow losses and yields from the other two subbasins with significant uncertainties suggest more complex coupled hydrologic systems in these reaches. Additional measurements of relative river–alluvial aquifer levels over the year, local groundwater withdrawals, and tributary inflows

Table 3. Mean, standard deviation for Q_{sub}, and net subbasin yield per km^2 or flow loss per m^2 of channel.

Subbasin	Q_{sub} (m^3/s)	*Net yield* $\times 10^{-4}$ ($m^3/s\text{-}km^2$)	*Net unit loss* $\times 10^{-7}$ ($m^3/s\text{-}m^2$)	Q_{sub} (m^3/s)	*Net yield* $\times 10^{4}$ ($m^3/s\text{-}km^2$)	*Net unit loss* $\times 10^{-7}$ ($m^3/s\text{-}m^2$)
		January 1977			*January 1979*	
Cr + WCC–Og	–0.48, 0.06		4.09, 0.50	0.57, 0.06		4.87, 0.55
Og + 2WCC–Ka	0.13, 0.15	0.18, 0.21		0.21, 0.13	0.28, 0.18	
Ka + WR–Oa	–0.20, 0.34		0.30, 0.50	–0.55, 0.31		0.81, 0.45
		February 1976			*November 1977*	
Cr + WCC–Og	0.49, 0.12	1.01, 0.25		–0.56, 0.05		4.84, 0.45
Og + 2WCC–Ka	9.05, 0.86	12.4, 1.18		1.94, 0.18	2.65, 0.25	
Ka + WR–Oa	0.53, 0.70	0.57, 0.75		1.81, 0.20	1.94, 0.21	

would reduce the uncertainty in the winter hydrologic balance and allow reliable estimates of the annual flow exchanges.

CONCLUSIONS

The semiarid White River basin is heterogeneous, with highly variable annual and winter average subbasin yields to the river caused by differences in soils and underlying strata. Winter is the season of minimum flows throughout the basin. The winter water balance is simplified because of the absence of quantities, such as evapotranspiration and water withdrawals, that are large in other seasons and have large uncertainties. We have developed a methodology for quantifying inflow to the river from a subbasin and the river–alluvial aquifer flow exchange by month through the winter. Important aspects of the method are a winter water balance equation with a river ice growth–melt term and a point estimate method that uses deterministic models with variable or uncertain parameters. The yield to the river from a subbasin is affected by precipitation, geology, and consumption. Trends in this parameter with time or between subbasins have direct water management implications.

The variable severity of the winters in our seven-year study period did not significantly affect the water balance. Water storage as ice is generally a dominant component of the water balance on the main-stem White River below Oglala, where the channel becomes wide. The large Crawford–Oglala and Kadoka–Oacoma subbasins on the main stem did not contribute flow to the river in most months of the study period. Even relatively mild winters did not produce inflows from these subbasins, unless a runoff event occurred. In contrast, the Oglala–Kadoka subbasin, situated between the others, consistently contributed flow to the river. The flow to this reach from three perennial creeks is the probable cause of this anomalous behavior. Very consistent monthly flow losses from the river at a sand–silt seepage velocity provide evidence of a predominantly perched river between Crawford and Oglala. Small, variable flow yields and losses suggest coupled hydrologic systems downstream, with the alluvial water table near (Oglala–Kadoka) or below (Kadoka–Oacoma) the river level during winter. These hypothetical hydrologic systems, based on the results of this study, are consistent with the field investigations of Rothrock (1942).

The mean, variance, and extremes obtained with the PEM for dependent variables such as water storage as ice and subbasin inflow allow definitive conclusions to be developed or identify the independent variables responsible for uncertainty in the results. Computation of air temperatures by subbasin instead of over the complete basin, and additional river width data to characterize a reach, would reduce the uncertainty in the present water balance. Improved estimates of the exchange between the river and the alluvial groundwater in a subbasin can be obtained by gaging all perennial creeks. A well-defined water balance that quantifies the winter river exchange with the alluvial aquifer in semiarid regions, together with measurements of the relative river–alluvial aquifer levels throughout the year, can provide reliable estimates of the annual flow exchange.

LITERATURE CITED

Abdulrazzak, M.J. and A.U. Sorman (1994) Transmission losses from ephemeral stream in arid regions. *Journal of Irrigation and Drainage, ASCE,* **120**(3): 669–675.

ASCE (1980) Characteristics of low flows. *Journal of the Hydraulics Division, ASCE,* **106**(HY5): 717–731.

Ashton, G.D. (Ed.) (1986) *River and Lake Ice Engineering.* Littleton, Colorado: Water Resources Publications, p. 42–48.

Ashton, G.D. (1989) Thin ice growth. *Water Resources Research,* **25**(3): 564–566.

Bear, J. (1972) *Dynamics of Fluids in Porous Media.* New York: American Elsevier, p. 132–136.

Ellis, M.J. and D.G. Adolphson (1971) Hydrogeology of the Pine Ridge Indian Reservation, South Dakota. Hydrologic Investigations Atlas HA-357. Washington, D.C.: U.S. Geological Survey.

Ellis, M.J., J.H. Ficken and D.G. Adolphson (1971) Hydrogeology of the Rosebud Indian Reservation, South Dakota. Hydrologic Investigations Atlas HA-355. Washington, D.C.: U.S. Geological Survey.

Gerard, R. (1981) Regional analysis of low flows: A cold region example. In *Proceedings of the 5th Canadian Hydrotechnical Conference, Fredericton, May 26–27,* Canadian Society for Civil Engineering, p. 95–112.

Harr, M.E. (1977) *Mechanics of Particulate Media: A Probabilistic Approach.* New York: McGraw-Hill, p. 474–512.

Kuusisto, E. (1986) Winter and summer low flows in Finland. In *Proceedings of the 6th International Northern Research Basins Symposium/Workshop, Houghton, Michigan, January 26–30,* p. 31–38.

Lacher, L.J., R. MacNish, T. Maddock III and L.G. Wilson (1994) Estimation of recharge along an effluent dominated reach of the Santa Cruz River, Tucson, Arizona, using a new approach to low-flow streamgaging (Abstract). 1994 Spring Meeting of the AGU, May 23–27, Baltimore, Maryland. Supplement to *EOS* **75**(16): 159.

Melloh, R.A. (1990) Analysis of winter low-flow rates in New Hampshire streams. USA Cold Regions Research and Engineering Laboratory, Special Report 90-26.

Mulherin, N.D., M.G. Ferrick, P.B. Weyrick, N.M. Perron and S.T. Hunnewell (1992) Development of a field technique for measuring fresh water ice density. In *Proceedings of the 49th Annual Eastern Snow Conference, Oswego, New York, June 3–4,* p. 203–212.

Prowse, T.D. (1990) Heat and mass balance of an ablating ice jam. *Canadian Journal of Civil Engineering,* **17**: 629–635.

Riggs, H.C. (1972) Low-flow investigations. *Techniques of Water-Resources Investigations of the U.S. Geological Survey,* Book 4, Chapter B1, Washington, D.C.

Rogers, J.D. and J.T. Armbruster (1990) Low flows and hydrologic droughts. In *Surface Water Hydrology* (M.G. Wolman and H.C. Riggs, Eds.), *The Geology of North America,* Vol. O-1, Geological Society of America.

Rosenblueth, E. (1975) Point estimates for probability moments. *Proceedings of the National Academy of Science U.S.A.,* **72**: 3812–3814.

Rothrock, E.P. (1942) A hydrologic study of the White River Valley, South Dakota. University of South Dakota, Vermillion, South Dakota, Geological Survey Rept. 41.

Sando, S.K. (1991) Estimation and characterization of the natural streamflow of the White River near the Nebraska–South Dakota state line. U.S. Geological Survey, Huron, South Dakota, Water-Resources Investigations Report 91-4096.

Schneider, W.J. (1965) Areal variability of low flows in a basin of diverse geologic units. *Water Resources Research,* **1**(4): 509–515.

Sophocleous, M. and S.P. Perkins (1993) Calibrated models as management tools for stream-aquifer systems: The case of central Kansas, USA. *Journal of Hydrology,* **152**: 31–56.

Wuebben, J.L., S.F. Daly, K.D. White, J.C. Tatinclaux, J.E. Zufelt, and J.J. Gagnon (1992) Ice impacts on flow along the Missouri River. In *Proceedings of the IAHR Ice Symposium, Banff, Alberta,* vol. 3, p. 1313–1322.

APPENDIX A: ROSENBLEUTH'S POINT ESTIMATE METHOD

The PEM of Rosenbleuth (1975) is described here in detail for a function of two random variables $y(x_1,x_2)$. The correlation coefficient of random variables x_1 and x_2 is defined as

$$\rho = \frac{Cov(x_1, x_2)}{S_{x_1}, S_{x_2}} \tag{A1}$$

where S_x is the standard deviation of x and $Cov(x_1,x_2)$ is the covariance. A correlation coefficient of 1 indicates a perfect linear correlation between the variables, and a coefficient of 0 indicates perfectly uncorrelated variables. Numerical experiments have shown that if the relationship between random variables over a limited range can be written as

$$x_1 = a x_2^b \, , \tag{A2}$$

then the magnitude of ρ will be approximately 1.

To simplify this discussion, we will assume that the distribution of each random variable is symmetric about the mean. The point estimates that represent the distribution of each random variable are then

$$P_{++} = P_{--} = \frac{1+\rho}{4} \tag{A3}$$
$$P_{+-} = P_{-+} = \frac{1-\rho}{4}$$

These point estimates are weighting factors that sum to 1 and, in the case of uncorrelated random variables, are each 1/4. The function $y(x_1,x_2)$ is evaluated at points that are a standard deviation from the mean of each random variable :

$$y_{++} = y(\bar{x}_1 + S_{x_1}, \bar{x}_2 + S_{x_2}) \tag{A4}$$
$$y_{+-} = y(\bar{x}_1 + S_{x_1}, \bar{x}_2 - S_{x_2})$$
$$y_{-+} = y(\bar{x}_1 - S_{x_1}, \bar{x}_2 + S_{x_2})$$
$$y_{--} = y(\bar{x}_1 - S_{x_1}, \bar{x}_2 - S_{x_2})$$

Then, the expected values or moments of the function can be obtained as

$$E(y^n) = P_{++} y_{++}^n + P_{+-} y_{+-}^n + P_{-+} y_{-+}^n + P_{--} y_{--}^n \, . \tag{A5}$$

The expected value or mean of y is found by setting $n = 1$. The variance of y is readily obtained as

$$V(y) = E(y^2) - [E(y)]^2 = S_y^2 \, . \tag{A6}$$

The minimum and maximum values of y fall outside the limits of the values obtained in eq A4. Rosenbleuth (1975) discussed the generalization of this method to functions of any number of random variables and random variables with asymmetric distributions.

In summary, the PEM provides estimates of the mean, variance, and limits of the distribution of a function of random variables. These four parameters uniquely specify a β-distribution, described by the general expression given by Harr (1977)

$$f(y) = \frac{1}{(b-a)B(\alpha+1,\beta+1)}\left(\frac{y-a}{b-a}\right)^{\alpha}\left(\frac{b-y}{b-a}\right)^{\beta} \quad \text{(A7)}$$

$$\alpha = \frac{(\bar{y}-a)^2}{S_y^2}(1-\tilde{y})-(1+\tilde{y})$$

$$\beta = \frac{\alpha+1}{\tilde{y}}-(\alpha+2)$$

$$\tilde{y} = \left(\frac{\bar{y}-a}{b-a}\right)$$

where a and b are the minimum and maximum values of y, and B is a beta function. When α and β have the same sign, the β-distribution is unimodal and bell-shaped. The coefficients of skewness and kurtosis of the β-distribution can be readily obtained. With this distribution, the probability of y in a given range can be determined.

Each model that was used to apply the PEM in this study has the same main program and a unique subroutine called EQN that contains the specific equation being solved. The main program performs all input and output functions and applies the PEM. The input variables and file structure are described in the comments contained in the main program, as well as the definitions of all variables used. In subroutine EQN each independent random variable has a number that gives its position in the array VAR. In the output file, the mean, coefficient of variation, standard deviation, and mean plus and minus standard deviation are given for each independent random variable by position in VAR. This list is followed by the mean, variance, standard deviation, coefficient of variation, and estimates of the maximum and minimum of the corresponding dependent variable. The EQN subroutines developed in this study are

- ROSICTHK, used to obtain the change in ice thickness from the air temperatures in the basin using the temperature index model,
- ROSQICE, used to obtain the discharge loss from the stream that goes into storage in a reach as ice,
- ROSQST, used to obtain the discharge loss to a stream to satisfy a change in water storage in a reach,
- ROSQLCL, used to obtain the discharge from the subbasin to a stream reach,
- ROSQYLD, used in cases with a negative local inflow from the subbasin to obtain the flow loss per square meter of wetted stream channel area.

Listings of these subroutines and of the main program follow.

```
      PROGRAM ROSICTHK
C
C   COMPUTE THE ICE THICKNESS FROM THE TEMPERATURE INDEX MODEL
C
      DIMENSION ID(515,9),RT(515),CVARP(9,7),CVARM(9,7)
      DIMENSION RHO(37),NRV1(37),NRV2(37),SIGN(37),RHT(37)
      DIMENSION XBAR(9),CV(9),SX(9),VARP(9),VARM(9),INDEX(9)
C
C------------------------------------------------------------------
C ROSENBLUETH'S FOSM APPROXIMATE PROCEDURE TO DETERMINE THE
C    MEAN AND STANDARD DEVIATION OF Y GIVEN THE FUNCTIONS  X(I), I=1,N.
C    THE SUBROUTINE EQN IS USED TO CALCULATE Y.
C
C    PROGRAMMED BY EARL EDRIS, WES PHONE 601-634-3378
C
C------------------------------------------------------------------
C
C   DEFINITIONS:
C
C     N       NUMBER OF RANDOM VARIABLES
C
C     XBAR    MEAN VALUE OF RANDOM VARIABLE
C
C     SX      STANDARD DEVAITON OF RANDOM VARIABLE
C
C     CV      COEFFICIENT OF VARIATION OF RANDOM VARIABLE
C
C     VARP    MEAN PLUS ONE STANDARD DEVIATION OF RANDOM VARIABLE
C
C     VARM    MEAN MINUS ONE STANDARD DEVIATION OF RANDOM VARIABLE
C
C    I,J,K,JJ INDEX VARIABLES FOR ARRAY SUBCRIPTING
C
C    TYPE    VARIABLE INDICATING TYPE OF DATA BEING INPUT
C            1 - ENTERING COEFFICIENT OF VARIATION
C            2 - ENTERING STANDARD DEVIATION
C
C    INDEX   ARRAY INDICATING HOW PLUS AND MINUS TERMS ARE OBTAINED
C            0 - CALCULATE VALUES AS MEAN PLUS OR MINUS SX
C            1 - ENTER VALUES FOR PLUS AND MINUS TERMS
C
C     NPM     NUMBER OF PLUS AND MINUS TERMS TO BE ENTERED FOR A RANDOM
C              VARIABLE
C
C    CVARP   ARRAY CONTAINING ENTERED PLUS VALUES
C
C    CVARM   ARRAY CONTAINING ENTERED MINUS VALUES
C
C     C      MAXIMUM NUMBER OF CORRELATION COEFFICIENTS
C
C     RHO     CORRELATION COEFFICIENT
C
C   NRV1,NRV2  RANDOM VARIABLES ASSOCIATED WITH RHO VALUE
C
C     TOTI    MAXIMUM NUMBER OF POSSIBLE COMBINATIONS OF PLUS AND
C            MINUS TERMS FOR N RANDOM VARIABLES
C
C     ID      ARRAY INDICATING WHEN PLUS AND MINUS TERMS SHOULD BE
C            USED FOR EACH COMBINATION GIVEN N RANDOM VARIABLES
C
C TT,TN,TOTN,P  VARIABLES OR INDEXES USED TO OBTAIN ID ARRAY
C
C    PP      ARRAY CONTAINING THE CALCULATED P VALUES
C
C   SUMP     SUMMATION OF P VALUES
C
C   SIGN     SIGN OF CORRELATION COEFFICIENTS BASED ON VALUES IN ID ARRAY
C
C   RHT      MULTIPLICATION OF SIGN AND RHO VALUES
C
C   RT       SUMMATION OF RHT VALUES
```

```
C
C  Y      DEPENDENT VARIABLE EXPRESSED AS A FUNCTION OF N RANDOM VAR.
C
C  Y2      SQUARE OF Y
C
C  YMAX     MAXIMUM VALUE CALCULATED FOR Y
C
C  YMIN    MINIMUM VALUE CALCULATED FOR Y
C
C  SUMY    SUMMATION OF THE QUANTITY Y MULTIPLIED BY PP
C
C  SUMY2   SUMMATION OF THE QUANTITY Y2 MULTIPLIED BY PP
C
C  VAR     VARIABLE ARRAY USED TO TRANSFER APPROPIATE CALCULATED
C         PLUS OR MINUS VALUES TO THE SUBROUTINE EQN TO CALCULATE Y
C
C  VR      VARIABLE ARRAY USED TO TRANSFER APPROPIATE INPUT PLUS
C         OR MINUS VALUES TO THE SUBROUTINE EQN TO CALCULATE Y
C
C  EY      EXPECTED VALUE OF Y
C
C  SY2     VARIANCE OF Y
C
C  SY      STANDARD DEVIATION OF Y
C
C  CVY     COEFFICIENT OF VARIATION OF Y
C
C---------------------------------------------------------------------
C
C  INPUT
C    NOTE: EQUATION TO CALCULATE Y MUST BE ENTERED INTO SUBROUTINE EQN
C          BEFORE COMPILING PROGRAM
C
C    FIRST LINE - N
C
C    SECOND LINE - XBAR,CV OR SX,TYPE,INDEX   FOR FIRST RANDOM VAR.
C
C              TYPE - 1 ENTER COEFFICIENT OF VARIATION
C                     2 ENTER STANDARD DEVIATION
C
C              INDEX - 0 CALCULATE PLUS AND MINUS TERMS
C                      1 ENTER PLUS AND MINUS TERMS
C
C       IF INDEX = 0 REPEAT SECOND LINE FOR NEXT RANDOM VARIABLE
C
C       IF INDEX = 1
C           THIRD LINE - NPM
C           FOURTH LINE - CVARP,CVARM CONTINUE FOR NPM LINES
C
C    AFTER ALL RANDOM VARIABLES ARE INPUT
C
C    NEXT LINE - NRHO
C
C      IF NRHO = 0 END OF INPUT
C
C      IF NRHO > 0 NEXT LINES CONTAIN - RHO,NRV1,NRV2
C
C
C---------------------------------------------------------------------
C
   COMMON /DAT/VAR(9),VR(9,7)
   REAL PP(515)
   INTEGER C,NPM(9)
C
   OPEN(UNIT=1,FILE='ROSINP')
   OPEN(UNIT=6,FILE='ROSOUT')
C
C  READ NUMBER OF RANDOM VARIABLES
C
   READ (1,*) N
```

```
C
C READ VARIABLE DATA
C
   DO 10 J=1,N
C
C  INPUT:  MEAN, VX OR STD,TYPE,INDEX
C  TYPE= 1 IF VX OR 2 IF STD
C  INDEX= 0 TO CALCU + - TERMS OR 1 TO ENTER VALUES
C
   READ (1,*) XBAR(J),CV(J),TYPE,INDEX(J)
   IF (TYPE.EQ.1) THEN
   SX(J)=XBAR(J)*CV(J)
   ELSE
   SX(J)=CV(J)
   CV(J)=SX(J)/XBAR(J)
   ENDIF
   IF (INDEX(J).EQ.1) THEN
C  INPUT:  # OF + OR - TERMS TO BE ENTERED
   READ(1,*) NPM(J)
   DO 12 K=1,NPM(J)
C  INPUT:  PLUS VALUE, MINUS VALUE


 12   READ(1,*) CVARP(J,K),CVARM(J,K)
   ELSE
   VARP(J)=XBAR(J)+SX(J)
   VARM(J)=XBAR(J)-SX(J)
   ENDIF
   WRITE(6,150) XBAR(J),CV(J),SX(J),VARP(J),VARM(J)
 10  CONTINUE
 150  FORMAT(5X,'XBAR=',F8.4,5X,'CV=',F5.3,5X,'SX=',F8.4,
   & /,10X,'PLUS=',F8.4,5X,'MINUS=',F8.4)
C
C RHO VALUES
C
C   MAX. NUMBER OF RHO VALUES
C
    C=(N*(N-1))/2
C
C   INITIALIZE RHO ARRAYS
C      RHO-ARRAY,NRV-NUMBER OF R. V. ASSOC. WITH RHO
C
    DO 14 I=1,C
    RHO(I)=0.0
    NRV1(I)=1
 14  NRV2(I)=1
C
C   READ NUMBER OF RHO VALUES TO BE INPUT
C
C  INPUT:  # OF RHO VALUES
    READ(1,*) NRHO
    IF (NRHO.EQ.0) GO TO 15
C
C   ERROR CHECK-MAKE SURE NRHO IS LESS THAN THE MAX NO. OF RHO VALUES
C
    IF(NRHO.GT.C)THEN
    NRHO=C
    WRITE(6,200) NRHO,C
     ENDIF
C
C  READ RHO VALUES AND ASSOC. RANDOM VARIABLES
C
    DO 20 J=1,NRHO
C  INPUT:  RHO, ASSOC R. V. 1 AND 2'
 18  READ(1,*) RHO(J),NRV1(J),NRV2(J)
C
C ERROR CHECK-MAKE SURE NRV1 DOES NOT EQUAL NRV2
    IF(NRV1(J).EQ.NRV2(J)) THEN
     WRITE(6,205) NRV1(J),NRV2(J)
   GO TO 18
    ENDIF
 20  CONTINUE
```

```
C
C  PRINT RHO ARRAY AND ASSOC. R. V.
C
      WRITE(6,120) (RHO(J),NRV1(J),NRV2(J),J=1,NRHO)
 120  FORMAT(10X,'RHO=',F7.3,5X,'NRV1=',I3,5X,'NRV2=',I3)
C
C  DEVELOP PLUS AND MINUS ARRAY
C
 15   CONTINUE
      K=-1
C
C  MAXIMUM NUMBER OF POSSIBLE COMBINATIONS
C
      TOTI=2**N
      DO 22 J=1,N
      TT=1
      TOTN=2**(J-1)
      P=TOTI/TOTN
      DO 24 JJ=1,P
      TN=TT+TOTN-1
      IF(TN.GT.TOTI) GO TO 22
      IF (K.EQ.-1) THEN
       DO 26 I=TT,TN
 26    ID(I,J)= K
       K=+1
      ELSE
       DO 28 I=TT,TN
 28    ID(I,J)= K
       K=-1
      ENDIF
      TT=TT+TOTN
 24   CONTINUE
 22   CONTINUE
C
C  DETERMINE SIGN OF RHO VALUES BASED ON THE SIGNS OF THE TWO NRV
C    VALUES
C
      SUMP=0.0
      DO 30 I=1,TOTI
      RT(I)=0.0
       DO 32 J=1,C
       SIGN(J)=ID(I,NRV1(J))*ID(I,NRV2(J))
       RHT(J)=SIGN(J)*RHO(J)
 32    RT(I)=RT(I)+RHT(J)
C
C  CALCULATE P VALUES
C
      PP(I)=(1/TOTI)*(1+RT(I))
C     WRITE(6,130) I,PP(I)
C130  FORMAT(10X,'P',I3,5X,F6.3)
      SUMP=SUMP+PP(I)
C
C  ENSURE P VALUES ARE BETWEEN 0 AND 1
C
      IF(PP(I).LT.0.0) THEN
       IF(PP(I).GT.1.0) THEN
      WRITE(6,210) PP(I)
      GO TO 99
       ENDIF
       ENDIF
 30   CONTINUE
       IF(SUMP.GT.1.001) THEN
        WRITE(6,215) SUMP
        GO TO 99
       ENDIF
C
C  CALCULATE Y MEAN AND STD OF Y
C
C     DO 34 J=1,N
C     IF (INDEX(J).EQ.1) THEN
C      DO 36 K=1,NPM(J)
```

```
C 36  VR(J,K)=(CVARP(J,K)+CVARM(J,K))/2
C     ELSE
C     VAR(J)=XBAR(J)
C     ENDIF
C 34  CONTINUE
C     CALL EQN(Y)
C     YMAX=Y
C     YMIN=Y
      YMIN=100000.
      SUMY=0.0
      SUMY2=0.0
      DO 40 I=1,TOTI
      DO 42 J=1,N
      IF (INDEX(J).EQ.1) THEN
      IF (ID(I,J).LT.0) THEN
       DO 44 K=1,NPM(J)
 44     VR(J,K) = CVARM(J,K)
      ENDIF
      IF (ID(I,J).GT.0) THEN
       DO 46 K=1,NPM(J)
 46     VR(J,K) = CVARP(J,K)
      ENDIF
      ELSE
      IF (ID(I,J).LT.0) VAR(J)=VARM(J)
      IF (ID(I,J).GT.0) VAR(J)=VARP(J)
      ENDIF
 42  CONTINUE
C
      CALL EQN (Y)
      IF (YMIN.GT.Y) YMIN=Y
      IF (YMAX.LT.Y) YMAX=Y
      Y2 = Y**2
      SUMY=SUMY+PP(I)*Y
      SUMY2=SUMY2+PP(I)*Y2
 40  CONTINUE
      EY=SUMY
      SY2=SUMY2-(EY**2)
      IF(SY2.LT.0.)SY2=0.
      SY=SQRT(SY2)
      CVY=SY/EY
C
      WRITE(6,160) EY,SY2,SY,CVY,YMAX,YMIN
 160  FORMAT(5X,'MEAN=',F10.4,5X,'VAR[X]=',F10.4,5X,'STD X=',
     & F10.4,/,5X,'COEFF OF VAR=',F5.3,5X,'MAX. Y=',F10.4,
     & 5X,'MIN. Y=',F10.4)
C
C  ERROR MESSAGES
C
 200  FORMAT(10X,'*** ERROR MESSAGE ***',/,10X,
     & 'ATTEMPTED TO ENTER',I3,' RHO VALUES WHEN ONLY',I3,
     & ' ARE POSSIBLE(PROGRAM CONT. W/ MAX ALLOWABLE)')
 205  FORMAT(10X,'*** ERROR MESSAGE ***',/,10X,
     & 'DIFFERENT RANDOM VARIABLES ARE NEEDED FOR RHO',/,
     & 'THE FOLLOWING VALUES WERE ENTERED',2I3,' (REENTER DATA)')
 210  FORMAT(10X,'*** ERROR MESSAGE ***',/,10X,
     & 'THE P(I) VALUE',F6.4,' IS NOT BETWEEN 0 AND 1.',
     & '(PROGRAM TERMINATED)')
 215  FORMAT(10X,'*** ERROR MESSAGE ***',/,10X,
     & 'THE SUM OF THE P VALUES IS',
     & F6.3,' WHICH IS ABOVE THE MAX VALUE OF 1.0.',
     & '(PROGRAM TERMINATED)')
C
 99  CLOSE(UNIT=1)
      CLOSE(UNIT=6)
      STOP 1111
      END
C
      SUBROUTINE EQN(Y)
C  THIS SUBROUTINE CALCULATES THE ICE THICKNESS AT THE END OF A MONTH
C   GIVEN THE THICKNESS AT THE START OF THE MONTH
      COMMON /DAT/VAR(9),VR(9,7)
C  VARIABLE 1=K, 2=RHO, 3=HIN, 4=HIA, 5=TAIR(C), 6=DT(DAYS), Y=HF (END OF MONTH
C      THICKNESS)
```

```
C
      IF(VAR(5).LT.0..OR.VAR(3).EQ.0.)GO TO 9
      Y=VAR(3)-VAR(4)*VAR(5)*VAR(6)*8.64/(VAR(2)*33.34)
      GO TO 11
C
  9   V14=VAR(1)/VAR(4)
      V142=V14*V14
      V32=VAR(3)*VAR(3)
      V12=2.*VAR(1)/VAR(2)
      Y=V142+V32+2.*V14*VAR(3)-VAR(6)*V12*VAR(5)*8.64/33.34
C
      IF(Y.GT.0.)GO TO 10
      Y=-SQRT(-Y)-V14
      GO TO 11
C
 10   Y=SQRT(Y)-V14
 11   CONTINUE
C
C     WRITE(6,200) Y
C200  FORMAT(15X,'Y=',F10.4)
      RETURN
      END
```

```
      PROGRAM ROSQICE
C
C     COMPUTE THE DISCHARGE (QICE) LOST TO ICE GROWTH IN A RIVER REACH
C
      SUBROUTINE EQN(Y)
      COMMON /DAT/VAR(9),VR(9,7)
C     VARIABLE 1=HF(ICE THICKNESS AT END OF MONTH, M), 2=HIN(ICE THICKNESS AT
C       START OF MONTH, M), 3=DT(LENGTH OF MONTH, DAYS), 4=BUPSTR(MEAN
C       STREAM WIDTH UPSTREAM PART OF REACH, FT), 5=BDNSTR(MEAN STREAM
C       WIDTH DOWNSTREAM PART OF REACH, FT), 6=DX(REACH LENGTH, MILES),
C       Y=QICE(M**3/S)
C
      BAVE=0.5*(VAR(4)+VAR(5))*0.3048
      DH=VAR(1)-VAR(2)
      Y=BAVE*DH*VAR(6)*1609.344/(VAR(3)*86400.)
C
C     WRITE(6,200) Y
  200 FORMAT(15X,'Y=',F10.4)
      RETURN
      END
```

```
      PROGRAM ROSQST
C
C     COMPUTE THE DISCHARGE (QST) LOST TO CHANNEL STORAGE IN A RIVER REACH
C
      SUBROUTINE EQN(Y)
      COMMON /DAT/VAR(9),VR(9,7)
C     VARIABLE 1=DT (LENGTH OF MONTH, DAYS), 2=BUPSTR (MEAN STREAM WIDTH
C     UPSTREAM PART OF REACH, FT), 3=BDNSTR (MEAN STREAM WIDTH DOWNSTREAM
C     PART OF REACH, FT), 4=DX (REACH LENGTH, KM), 5=DY (CHANGE IN MEAN DEPTH
C     UPSTREAM PART OF REACH, FT), 6=DY(CHANGE IN MEAN DEPTH DOWNSTREAM
C     PART OF REACH, FT), Y=QST(M**3/S)
C
      Y=VAR(4)*1000/(86400*VAR(1))*((VAR(2)*VAR(5)+VAR(3)*VAR(6))*
     &.09290304)*.5
C
C     WRITE(6,200) Y
  200 FORMAT(15X,'Y=',F10.4)
      RETURN
      END
```

```
      PROGRAM ROSQLCL
C
C   COMPUTE THE DISCHARGE EXCHANGE WITH THE LOCAL BASIN (QLOCAL)
C    POSITIVE--FLOW FROM SUBBASIN TO RIVER;  NEGATIVE--FLOW LOSS FROM THE
C    RIVER TO THE ALLUVIAL AQUIFER
C
      SUBROUTINE EQN(Y)
      COMMON /DAT/VAR(9),VR(9,7)
C  VARIABLE 1=QST(FLOW GOING INTO CHANNEL STORAGE, M**3/S), 2=QOUT
C  (MEASURED OUTFLOW FROM A REACH, FT**3/S), 3=QIN (MEASURED INFLOW TO A
C  REACH, FT**3/S), 4=QICE (FLOW GOING INTO STORAGE AS ICE, M**3/S),
C  Y=QLOCAL (M**3/S)
C
      Y=(VAR(2)-VAR(3))*0.02831685+VAR(1)+VAR(4)
C
C    WRITE(6,200) Y
 200  FORMAT(15X,'Y=',F10.4)
      RETURN
      END
```

```
      PROGRAM ROSQYLD
C
C  FOR CASES WITH NEGATIVE QLOCAL, COMPUTE THE FLOW LOSS FROM THE
C  STREAM PER M**2 OF WETTED AREA (QYLD)
      IMPLICIT REAL*8(A-H,O-Z)
C
      SUBROUTINE EQN(Y)
      IMPLICIT REAL*8(A-H,O-Z)
      COMMON /DAT/VAR(9),VR(9,7)
C  VARIABLE 1=QLCL (FLOW FROM RIVER TO ALLUVIAL AQUIFER, M**3/S), 2=BUPSTR
C  (MEAN STREAM WIDTH UPSTREAM PART OF REACH, FT), 3=BDNSTR (MEAN
C  STREAM WIDTH DOWNSTREAM PART OF REACH, FT), 4=DX (REACH LENGTH, KM)
C  Y=QYLD(M**3/S/M**2)
C
      Y=VAR(1)/((.5*(VAR(2)+VAR(3))*.3048)*1000.*VAR(4))
C  WRITE(6,200) Y
 200  FORMAT(15X,'Y=',D12.5)
      RETURN
      END
```

REPORT DOCUMENTATION PAGE

Form Approved
OMB No. 0704-0188

1. AGENCY USE ONLY (Leave blank)	2. REPORT DATE	3. REPORT TYPE AND DATES COVERED
	July 1995	

4. TITLE AND SUBTITLE

The Winter Low-Flow Balance of the Semiarid White River, Nebraska and South Dakota

5. FUNDING NUMBERS

6. AUTHORS

Michael G. Ferrick, Nathan D. Mulherin and Darryl J. Calkins

7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES)

U.S. Army Cold Regions Research and Engineering Laboratory
72 Lyme Road
Hanover, New Hampshire 03755-1290

8. PERFORMING ORGANIZATION REPORT NUMBER

CRREL 95-15

9. SPONSORING/MONITORING AGENCY NAME(S) AND ADDRESS(ES)

Office of the Chief of Engineers
Washington, D.C. 20314-1000

10. SPONSORING/MONITORING AGENCY REPORT NUMBER

11. SUPPLEMENTARY NOTES

12a. DISTRIBUTION/AVAILABILITY STATEMENT

Approved for public release; distribution is unlimited.

Available from NTIS, Springfield, Virginia 22161

12b. DISTRIBUTION CODE

13. ABSTRACT (*Maximum 200 words*)

Low-flow studies are needed to quantify the effects of water consumption on stream flow, water quality, groundwater resources, and contaminant transport. The low-flow water balance of a river in a cold region is simplified in winter because evapotranspiration is negligible, irrigation water withdrawals and diversions are halted, and precipitation occurs largely as snow, minimizing the spatial and temporal variability of runoff. We investigated the monthly low-flow water balance of White River (Neb. and S. Dak.) reaches over seven consecutive winters. Water going into or out of storage as ice or melt, obtained with an air temperature index model, can be a dominant component of the water balance. The point estimate method is used to account for parameter uncertainty and variability, providing the mean, variance, and limits of dependent variables such as water storage as ice and inflow from a subbasin. Negative surface water yield from several-thousand-square-kilometer subbasins occurred regularly through the period, indicating a significant flow from the river to the alluvial aquifers. The winter water balance results suggest either a perched river or a coupled surface water–groundwater hydrologic system in particular subbasins, consistent with the field investigations of Rothrock (1942). The winter flow exchange between the surface and subsurface can be used to estimate the annual exchange for both hydrologic conditions.

14. SUBJECT TERMS

		Semiarid basin
Low flow	Point estimate method	Streamflow
Perched river	River ice growth	Winter water balance

15. NUMBER OF PAGES
31

16. PRICE CODE

17. SECURITY CLASSIFICATION OF REPORT	18. SECURITY CLASSIFICATION OF THIS PAGE	19. SECURITY CLASSIFICATION OF ABSTRACT	20. LIMITATION OF ABSTRACT
UNCLASSIFIED	UNCLASSIFIED	UNCLASSIFIED	UL